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Cumulus Convection and Larger Scale Circulations
I. Broadscale and Mesoscale Considerations

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ABSTRACT—This paper (part I) is a discussion of the magnitude and implication of the vertical circulation patterns of the summertime tropical atmosphere as derived from synoptic scale considerations. Part I is compared to the vertical circulation patterns derived from cumulus scale considerations as discussed by López in part II, a companion paper. From the synoptic scale considerations, we show that a very significant subsynoptic or local vertical motion is occurring within the cloud regions of the Tropics. This mass-cancelling local up- and-down circulation is not resolved by the mean or synoptic scale flow patterns. The magnitude of this local or up- and-down vertical circulation can be estimated from cloud-cluster scale (approx. 4°) mass, vapor, energy, and rainfall-evaporation budgets. Results are closely comparable to those obtained by López from an independent small-scale approach through modelling of individual cumulus elements. This local vertical circulation is shown to be fundamental for the mass, vapor, and energy balances of the tropical atmosphere. Other discussions of the characteristics of the cumulus convective atmosphere are included.

1. INTRODUCTION

The manner by which cumulus clouds and the broader scale flow patterns interact is not well understood at this time. It is important that we consider the physics of this interaction problem. Many meteorologists believe this to be a fundamental requirement to improved understanding and prediction of large-scale atmospheric flow patterns. The instigation of the GARP [1] Atlantic Tropical Experiment has been in response to this requirement for more physical understanding of the cumulus processes.

The author believes that a significant expansion of our knowledge on this problem is possible from the meteorological data already on hand—if we organize our various facets of information in a judicious way. This is the purpose of the following discussion.

This research includes the required mass, water vapor, and energy budgets of the summer, oceanic trade wind-equatorial trough belt from about 5° to 25° latitude. These budgets are obtained by resolving into a mutually consistent pattern the available broad-scale (meso, synoptic, and zonal) observational knowledge on the mass, water vapor, and energy of this belt.

In the paper, we show that the accomplishment of these balances requires a substantial local or subsynoptic scale up-and-down vertical circulation with condensation and re-evaporation rates much larger than the observed rainfall-evaporation.[2] In part I, we specify the magnitude of this vertical circulation and water-vapor recycling and include a discussion of the resulting energy requirements. These mass-vapor-energy budgets that are derived from broad-scale considerations for the cloud cluster are compared to the same budgets in part II by López (1973*c*) that are obtained from incorporation of individual convective elements of his cumulus life-cycle model. We will show that both approaches, one from the large scale going downward in scale (broad-scale approach) and the other using the individual convective elements and going upward in scale (cumulus scale approach), do mesh with near identical mass-vapor-energy budget results. The meshing of these independent approaches from different scales of consideration lends confidence to the results.

2. DATA SOURCES

To deal with the tropical belt in a realistic way, one must obtain representative information on such aspects as the typical tropical belt lapse rate conditions, vapor contents, divergences, and shears associated with the satellite-observed tropical cloud clusters, other variable cloud areas, and clear regions. To accomplish this, the author performed extensive radiosonde data composite analysis of satellite-observed cloud and clear areas in the Western North Pacific Ocean and the West Indies. These regions have the only oceanic radiosonde networks from which associated wind-temperature-moisture information could be obtained.

Figures 1 and 2 show the locations of the radiosonde networks that were used to composite data around the digitized satellite mesoscale cloud clusters, other variable cloud regions, and clear regions for the three summer seasons of 1967–69. The methods of compositing and a discussion of the data limitations and inaccuracies have been made in an earlier report by Williams and Gray (1973) to which the reader is referred for more information on the data-reduction procedures. Figure 3 shows typical

[1] Global Atmospheric Research Program

[2] In this paper, we define local circulation as subsynoptic scale motion; whereas mean motion is used synonymously with synoptic scale flow.

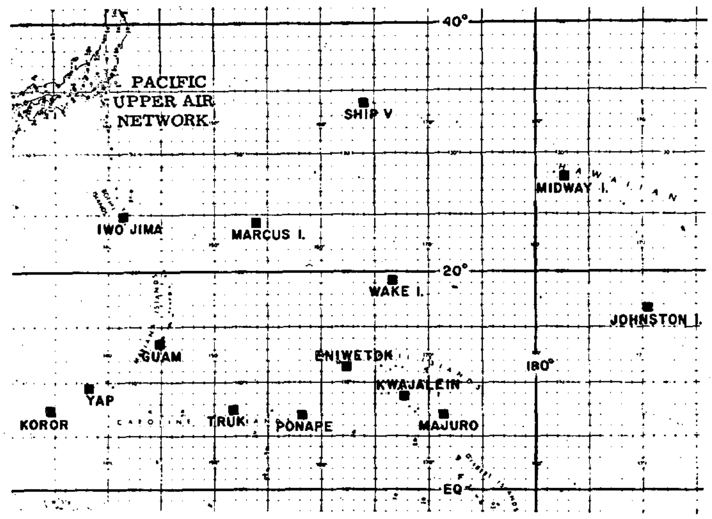

FIGURE 1.—Radiosonde network in the Western North Pacific Ocean.

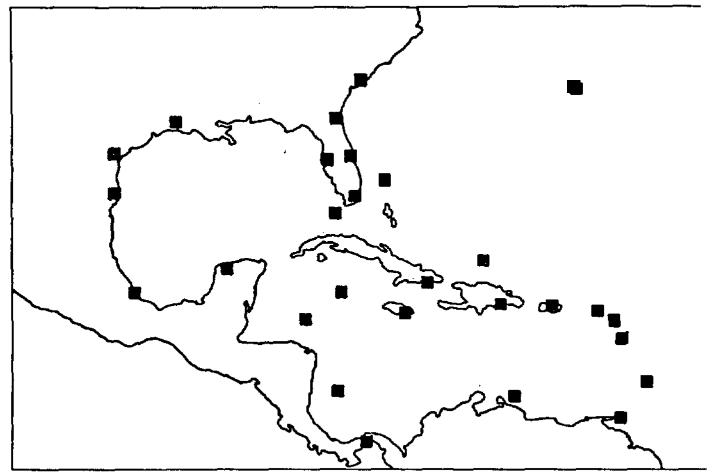

FIGURE 2.—Radiosonde network in the West Indies.

views of these satellite-observed tropical clusters and clear regions.

Altogether, there were 557 clusters and 223 clear areas in the Western North Pacific and 539 clusters and 212 clear areas in the West Indies networks included in the data composites.

3. ENTIRE TROPICAL BELT RAINFALL, WATER VAPOR, AND ENERGY BUDGET CONSIDERATIONS

Tropospheric conditions in the whole Summer tropical belt are considered. This belt includes the trade wind and equatorial trough regions from approximately 5° to 25° latitude. Oort and Rasmusson (1971) have shown (from data derived from the MIT[3] Department of Meteorology rawinsonde tapes) that this tropical area is (from the standpoint of mass, vapor, and energy budgets) a self-contained region. Their data indicate that the net tropospheric energy divergences in the summertime latitudes from 5° to 25° latitude are much smaller than the net tropospheric radiational cooling. They show that, for mean tropospheric conditions over this region, vertical motion averages only a few millibars per day. Meridionally induced advective cooling-warming rates (averaged through the troposphere) are less than 0.1°C/day. In comparison, the lower tropospheric up-moist vertical motion needed to accomplish the required rainfall and the down-dry vertical motion needed to balance the net radiation cooling must be of the order of 100 mb/day. Evaporation-rainfall must average about 0.5 cm/day or about 1.3°C latent heat equivalent for the whole troposphere. Thus in comparison with net tropospheric radiational losses, the summertime meridional fluxes in the broad tropical belt where cloud clusters exist can be largely neglected. The summertime tropical region from 5° to 25°N must closely meet its own energy budget requirements. The longitudinal or Walker circulations allow for some individual longitude budget imbalances, but these largely cancel in the global latitudinal average. Figure 4 shows that the net meridional divergence of energy is much smaller than the net radiation loss. (Note the lack of any appreciable energy convergence in the Equator to 20° latitude belt.) We assume that this broad-scale estimate also applies to the summertime oceanic regions by themselves.

Over the tropical oceans, the ratio of sensible to latent heat transport is estimated to be small (Sellers 1965). If we assume negligible sensible heat transport, then the tropospheric radiation losses of about 1.3° C/day (table 1) must be balanced by an average summertime rainfall-evaporation rate of about 0.5 cm/day because of the self-contained nature of the tropical belt. Previous estimates (Budyko 1956) indicate that this value is a reasonable estimate of the summertime tropical rainfall-evaporation rate. Evaporation of 0.5 cm/day will be assumed everywhere. For simplicity, the net tropospheric radiation cooling and the energy of evaporation will be assumed constant and everywhere balanced as in figure 5.

Although evaporation is rather uniformly distributed, rainfall is typically concentrated in mesoscale cloud clusters comprising only 15–20 percent of the area of the tropical belt. Here, the total tropical belt evaporation falls out as rain in average amounts of 2–3 cm/day (Williams and Gray 1973). Other regions possess cloudiness but with negligible amounts of precipitation. The dynamics of the rain, cloud, and clear regions must be quite different. Each region has its own distinctive mass, vapor, and energy budget, which should be treated individually.

Simplified Division of Tropical Belt

To expedite understanding of the tropical belt, it is divided into three meteorological regions. These are:

1. Cloud cluster regions taking up about 20 percent of the area of the tropical belt and having average rainfall of 2.5 cm/day.
2. Variable cloud regions taking up about 40 percent of the area of the tropical belt with scattered to broken cloud conditions but no applicable rainfall. The average temperature and dew-point conditions of this region were obtained from compositing the regions around the cloud cluster as seen in figure 10.
3. Clear regions taking up about 40 percent of the area of the tropical belt.

[3] Massachusetts Institute of Technology

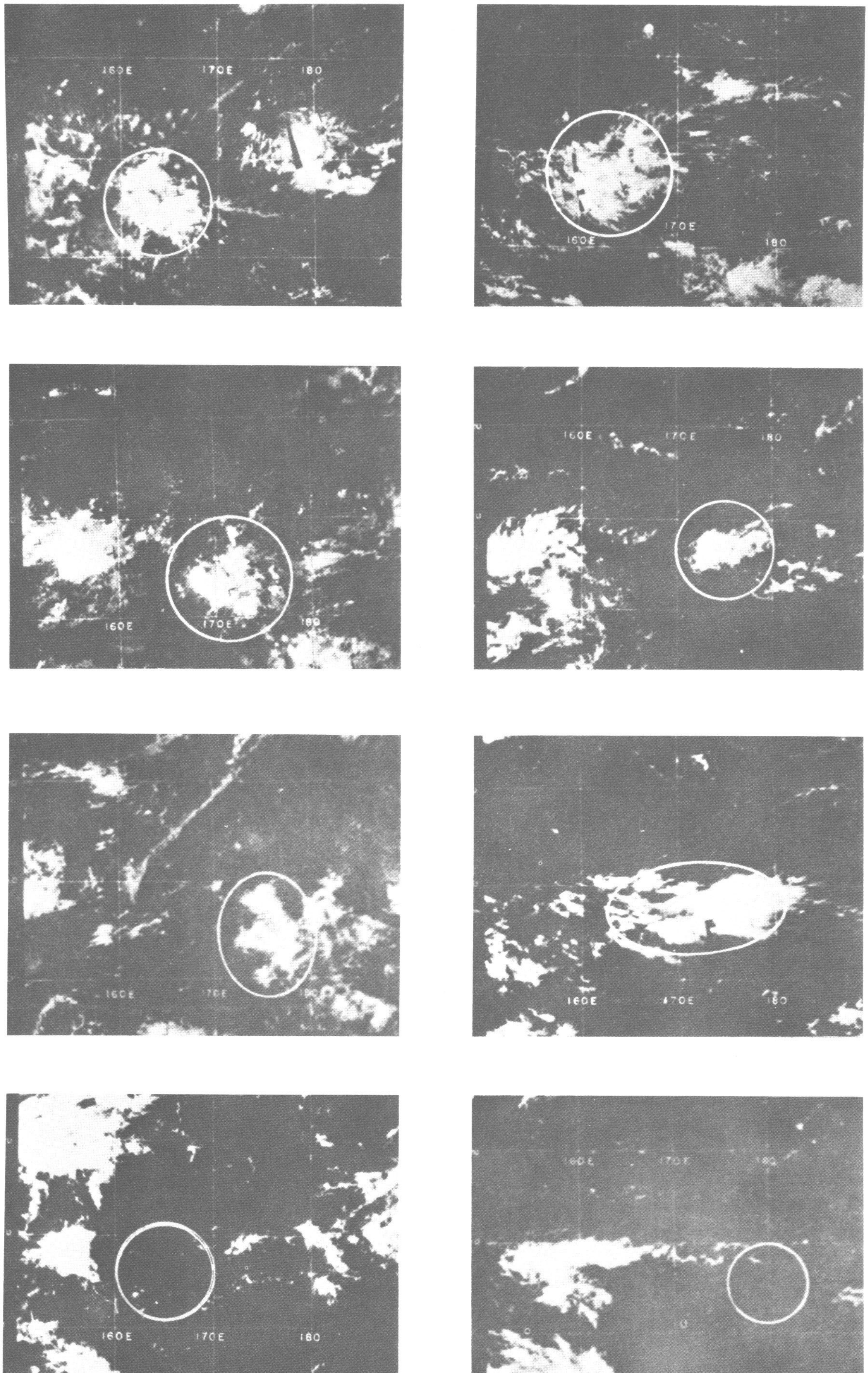

FIGURE 3.—Typical portrayal of satellite-observed cluster and clear regions.

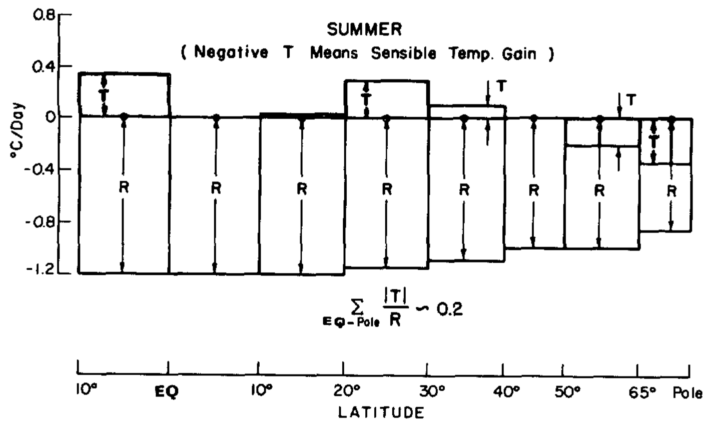

FIGURE 4.—Comparison of the mean tropospheric temperature change due to all meridional energy convergence (data from Oort and Rasmusson 1971) sources (T) by wind systems versus net tropospheric radiation (data from sources in table 1) loss (R) in the layer 950–150 mb for summer.

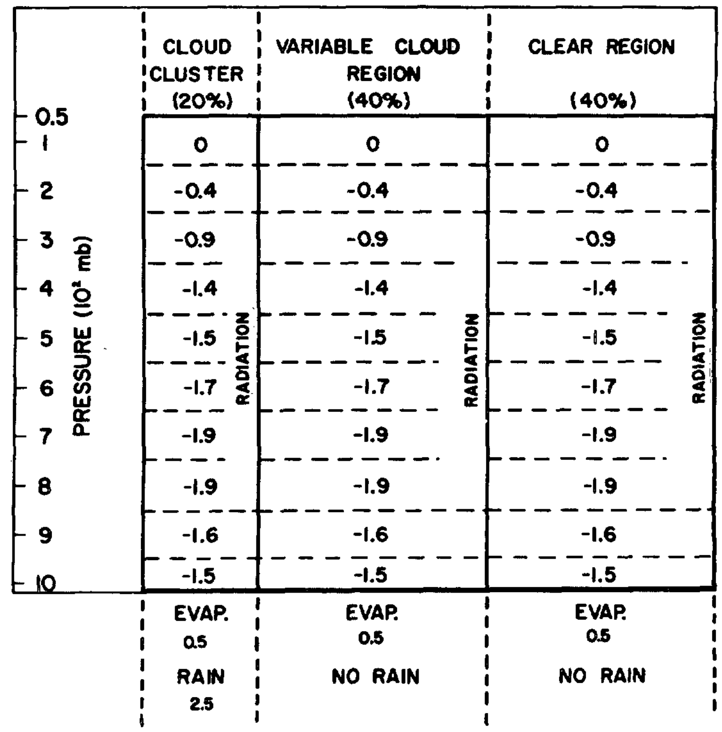

FIGURE 5.—Assumed net radiation cooling (°c/day), evaporation ($g \cdot cm^{-2} \cdot day^{-1}$), and rain ($g \cdot cm^{-2} \cdot day^{-1}$) in the three tropical regions.

TABLE 1.—*Estimates of Equator to 30°N net radiation cooling (°C/day) in the layer from 1000 to 150 mb. This tropical region has little seasonal variation.*

		EQ-30°
Model determinations	London (1957)	1. 12
	Davis (1963)	1. 11
	Rogers (1967)	1. 18
	Dopplick (1970)	1. 13
	Average	1. 14
Measurements (with Davis short-wave values)	Cox and Suomi (1969)	1. 37
Measurements	Vonder Haar (1971)	1. 26 (EQ–20°)

Figures 6 and 7 show individual region temperature and dew-point values obtained from the rawinsonde compositing procedure. Note that no significant temperature differences exist between the three regions; therefore, horizontal temperature advection must be near zero. The three regions are primarily contrasted by their middle and lower middle tropospheric moisture differences.

Figure 8 portrays these three typical regions in idealized cross-section form. Observational research of Chang (1970), Frank (1970, 1971), Wallace (1970), Hayden (1970), and Martin and Suomi (1972) gives general support to the above area percentage classification.

Each region now will be discussed in more detail. We will first establish the typical area divergences and mean or synoptic scale vertical motion profiles.

Clear area region. The clear area is the simplest region. The net radiation cooling as portrayed in figure 5 is assumed to be exactly balanced by a sinking compressional warming of about 25–30 mb/day. Composites of the winds around the clear areas independently verify this magnitude of sinking motion. The decrease of water vapor content as a result of this sinking motion is made up at higher levels by vapor advection from the variable cloud region.

Cloud cluster region. The composited cloud cluster data of the Western North Pacific (Williams and Gray 1973) and the West Indies (Gray and Ruprecht 1974) show that the mesoscale cloud cluster of about 4° latitude width has a deep, nearly constant convergence up to about 400 mb and a strong concentrated divergent outflow layer at 200 mb. Reed and Recker (1971) and Yanai et al. (1973) have obtained or implied a similar mass composite profile for a number of easterly waves and/or cluster systems in the central Pacific. Other information from tropical disturbances indicate a similar vertical convergence arrangement. This cluster convergence pattern up to 400 mb is responsible for an average 4° cluster water-vapor inflow of about 1.5–2.0 $gm \cdot cm^{-2} \cdot day^{-1}$ (Williams and Gray 1973). This is necessary for a mean cluster rainfall of 2.0–2.5 cm/day if 0.5 cm/day evaporation is occurring underneath the cluster. These cluster rainfall estimates have also been independently verified from Pacific atoll and island rainfall data composited with respect to the clusters. The cloud clusters and the clear areas thus are regions that import water vapor.

We will show that the mean or synoptic scale upward circulation at low levels as determined by the cluster mass convergence profile falls far short of accounting for the observed upward vapor transport and rainfall. It is necessary to hypothesize an additional local or subsynoptic up-moist and down-dry circulation to accomplish all of the necessary upward vapor transport to explain the rainfall.

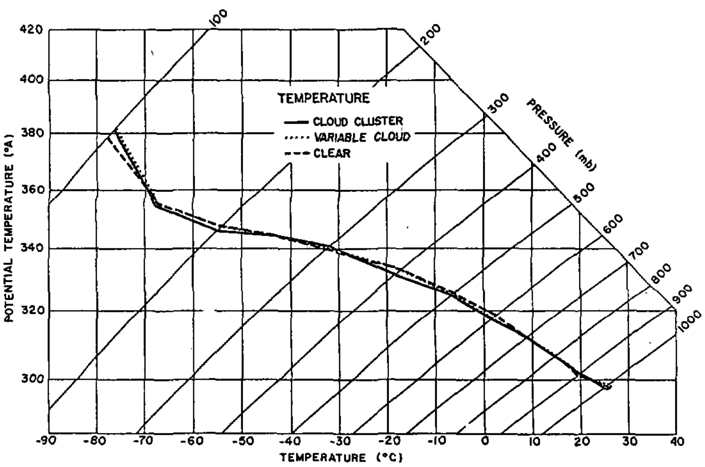

FIGURE 6.—Composite temperature soundings for the three regions.

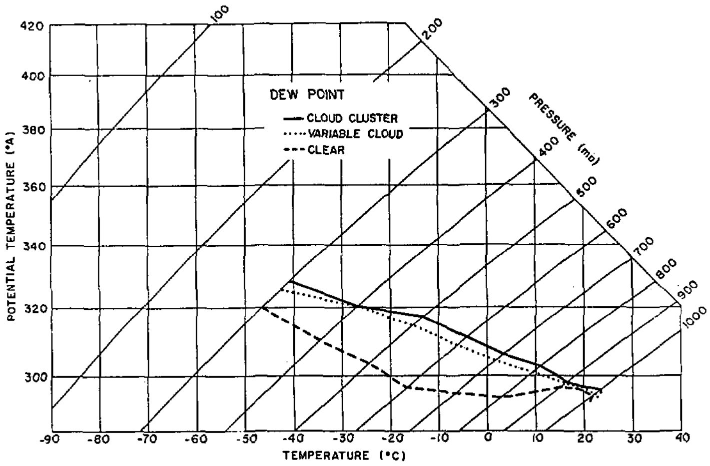

FIGURE 7.—Composite dew-point soundings for the three regions.

Variable cloud region. Given the average vertical motion pattern of the clear and cluster regions, we solved for the residual, which represents the mean vertical motion of the variable cloud region. This motion is downward everywhere except in the boundary layer.

Since the variable cloud region must export water vapor to both the cluster and the clear regions, there must be (as with the cluster region) a large local or subsynoptic upward vapor transport against the downward vapor transport of the mean or synoptic scale circulation. This means that the upward transports of vapor by clouds are greater than the downward transports of air between the clouds and that the reverse is true for mass.

Figure 9 portrays the mean vertical motion for the three regions in millibars per day. The characteristics of these local or subsynoptic vertical circulations now will be discussed.

4. PROCEDURES FOR DETERMINING SUBSYNOPTIC VERTICAL CIRCULATION

With the definition of the three distinctive tropical regions and the general discussion of the self-contained nature (lack of significant outside meridional transports) of the tropical belt in summer, we now are in a position to estimate the complete mass and vapor budgets for each region separately. These budgets require a steady-state assumption for each region.

After the whole oceanic tropical belt budgets have been made, the individual region budgets are determined in the following order for:

1. Divergence and mean vertical motion.
2. Water vapor.
3. Local or subsynoptic vertical circulation.

The computational steps are made in sequence and are outlined by the flow diagram shown in figure 10. Following this diagram, we made the following assumptions and computations to determine the real up-and-down vertical motions in each region.

1. Assumptions for the whole tropical belt budget are:
   a. The entire oceanic tropical region of summer is a self-contained energy system.
   b. The entire region net radiational cooling is balanced by condensation warming.
   c. The average precipitation and surface evaporation are balanced for the entire tropical belt.
2. The computational steps for the individual region mass budgets are:
   a. Determine the mean *cluster* divergence by compositing rawinsonde data (Williams and Gray 1973).
   b. Determine the mean sinking in the *clear* regions by assuming it is enough to just balance the radiational cooling.
   c. Solve for the *variable cloud* region mean vertical motion and mass profile as a residual of the cluster and clear regions.
3. The computational steps for the individual region vapor budgets are:
   a. Obtain the cluster rainfall and vapor convergence from observations. From this, determine the vertical vapor transport requirements.
   b. Determine the clear region vapor convergence from its sinking drying.
   c. Solve for the variable cloud vapor requirements as a residual of the other two regions.
4. The computational steps for the required local (subsynoptic) up-and-down vertical circulation are:
   a. Determine the cluster subsynoptic vertical circulation from required vertical vapor transport, which synoptic vertical circulation does not accomplish.
   b. Follow the same procedure for *variable cloud* and clear regions.

Individual Region Divergence and Mean Vertical Motion

The method of determining the individual region divergence and the mean vertical motion from rawinsonde data compositing has been discussed previously by Williams and Gray (1973). Figures 11 and 12 show the divergence and mean vertical motion profiles for each region.

Determination of Individual Region Water Vapor Budgets

Water-vapor budgets for the three regions were made in a way similar to those for the mass budgets. The cluster vapor convergence vertical profile was determined from

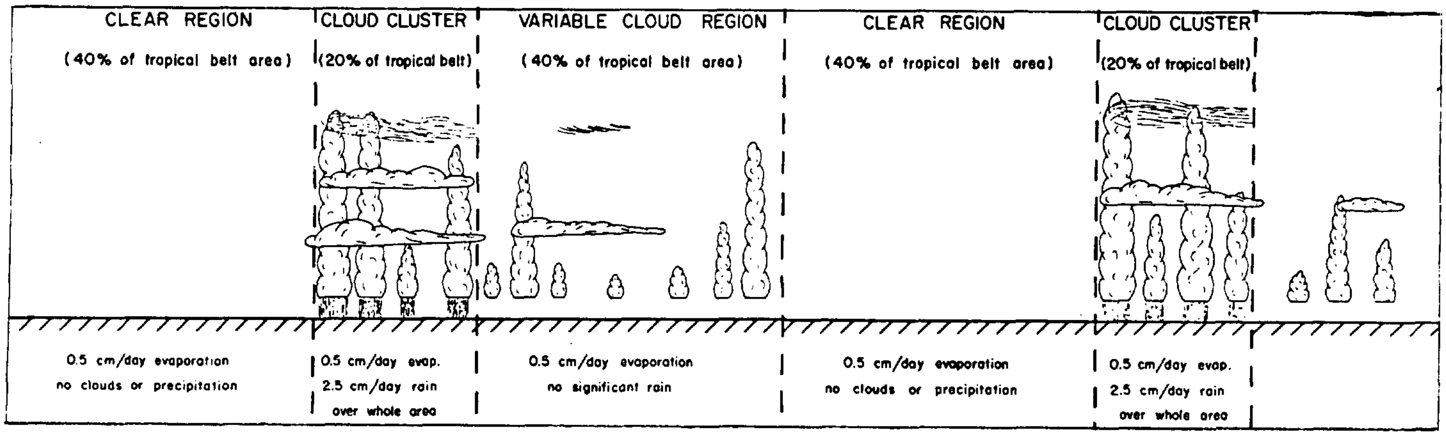

FIGURE 8.—Schematic of cloud-rain and clear areas in the Tropics during summer.

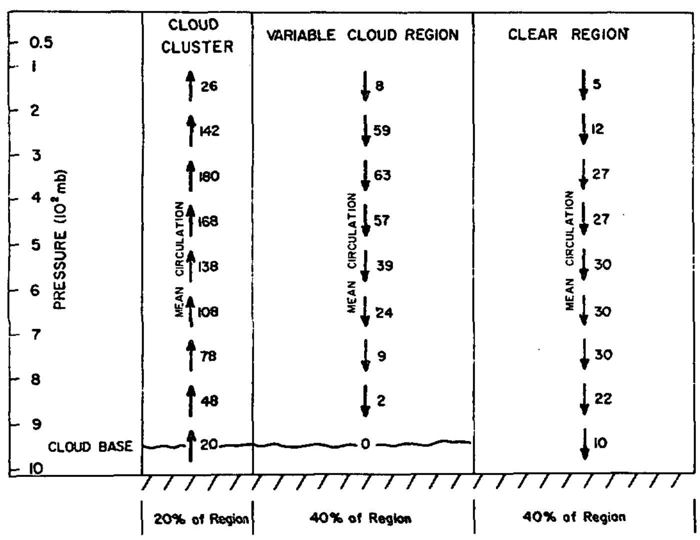

FIGURE 9.—Mean vertical motion pattern (mb/day) in the various regions.

the composite radiosonde information. The clear region vapor budget was specified by the rate of drying of each region due to the mean sinking motion. The variable cloud region vapor profile was obtained as a residual of the cluster and clear regions.

Figure 13 portrays the vertical distribution of water-vapor transport in vertical layers in and out of the cluster, variable cloud, and clear regions. The cluster region has a net water-vapor import of 2 $g \cdot cm^{-2} \cdot day^{-1}$. The 0.5 $g \cdot cm^{-2} \cdot day^{-1}$ of evaporation allows for an average daily rainfall of 2.5 $g \cdot cm^{-2}$. The clear region and the variable cloud region are net sources of water vapor. The clear regions import vapor at upper levels to balance their sinking drying motion of 0.36 $g \cdot cm^{-2} \cdot day^{-1}$, the boundary layer accumulates vapor at a daily rate of 0.86 $g \cdot cm^{-2}$. The clear areas thus are source regions of water vapor. This is primarily because of their boundary layer divergence. At upper levels, they import vapor.

The variable cloud region exports the vapor it receives from its evaporation (i.e., 0.5 $g \cdot cm^{-2} \cdot day^{-1}$). Above the boundary layer, it exports 0.86 $g \cdot cm^{-2} \cdot day^{-1}$, which can only occur if the vapor is carried to upper levels by a substantial subsynoptic circulation. Below cloud base, the variable cloud region imports 0.36 $g \cdot cm^{-2} \cdot day^{-1}$.

Since the rainfall of the cloud cluster is approximately four times the amount of the evaporation beneath the cluster, it is obvious that the cloud cluster must import a substantial amount of water vapor. This vapor import is accomplished by two processes:

1. Convergence of vapor from the surface to 400 mb owing to the mean convergence profile as seen in figure 11. This process accounts for most of the vapor convergence into the cluster.
2. Eddy advection of vapor into the cluster beyond that specified by the divergence. This occurs primarily in the lowest 50–150 mb and is analogous to the extra vapor advection into squall lines when they overtake lower level air, as discussed by Newton (1963). Typically, dryer downdraft air leaves the cluster on the upwind or rear side while moister lower level air enters the cluster from the downwind or forward side. This brings about a net moisture advection into the cluster. Zipser (1969, 1972) has discussed the large drying from downdrafts, which occurs in clusters. Cluster surface winds represent only half the winds at the 800- to 900-mb level.

An estimate of the vertical distribution of vapor advection into the cluster below any level p_i from these two processes thus can be obtained from the equation

$$\text{vapor advection into cluster} = \overbrace{\int_{sfc}^{p_i} \nabla_2 \cdot q \mathbf{V}_i \frac{\delta p}{g}}^{\text{vapor transfer by convergence}} + \overbrace{\int_{sfc}^{p_i} \Delta q_i \frac{|V_i - V_c|}{D} \frac{\delta p}{g}}^{\text{extra vapor into cluster at lower levels by relative motion}} \quad (1)$$

where p_i is the pressure at any level; Δq_i, the vapor difference between air entering and leaving the cluster at any level (typically 1–5 g/kg at lower levels); V_i, the wind at any level in the cluster; V_c, the cluster velocity; and D, the width of the cluster, taken as 4° latitude.

The main contribution of the second term on the right of eq (1) comes in the boundary layer. Here, the speeds are substantially less than the cluster velocity. The cluster vapor advection determined from eq (1) is shown for the cluster on the left portion of figure 13. Note that half the vapor advection comes in the boundary layer (surface to 950 mb).

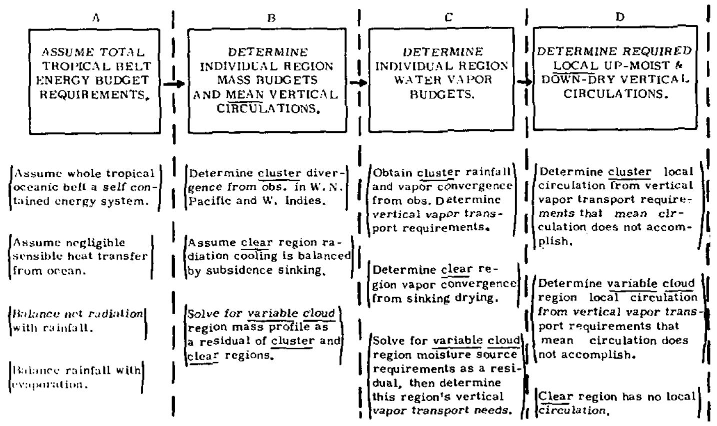

FIGURE 10.—Flow diagram of the computational method for the local circulation.

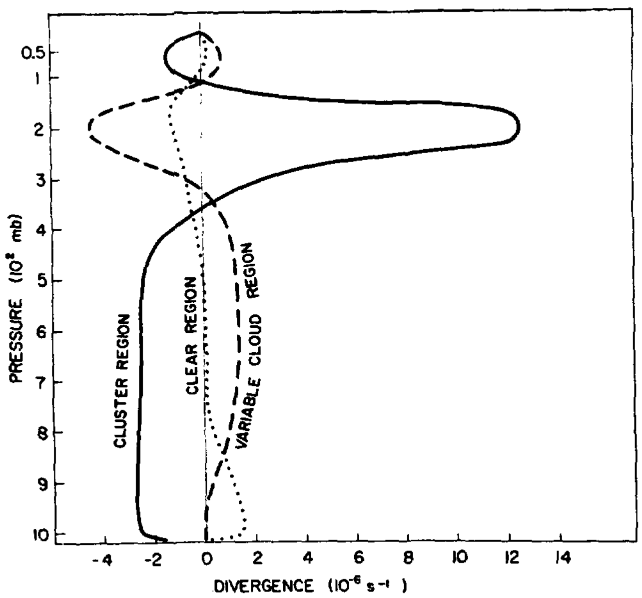

FIGURE 11.—Divergence profiles for each region.

One should realize that the 2.5 cm/day rainfall is not occurring everywhere within the cluster but is probably concentrated in local rain areas (some along squall lines) taking up only 10 percent or less of the 4° wide cluster. In addition, it is likely that only about 10–20 percent (or 0.2–0.4 percent of the entire tropical belt) of the rain areas have active towering cumulus or cumulonimbus updrafts in operation. A similar scaling of such aspects as the relative areas of the updraft, rain regions, and weather systems, has been presented by Riehl and Malkus (1958). Even though the convective activity is highly concentrated within the cluster, the lack of appreciable broadscale temperature and water-vapor gradients across the cluster allows for a cluster area average treatment of the mass, vapor, and energy budgets.

Calculation of Required Subsynoptic Vertical Circulation

The required subsynoptic circulation in the cluster and variable cloud regions can be readily determined from the

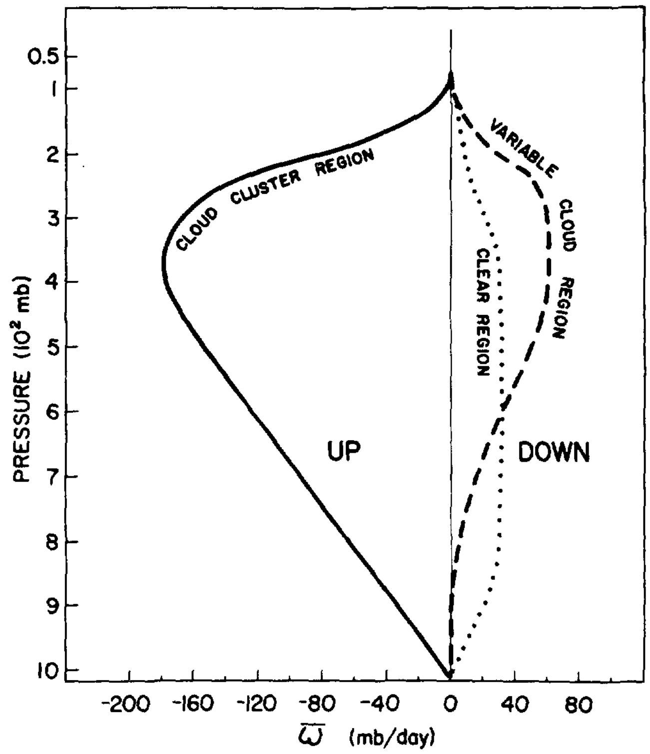

FIGURE 12.—Mean vertical motion for each region.

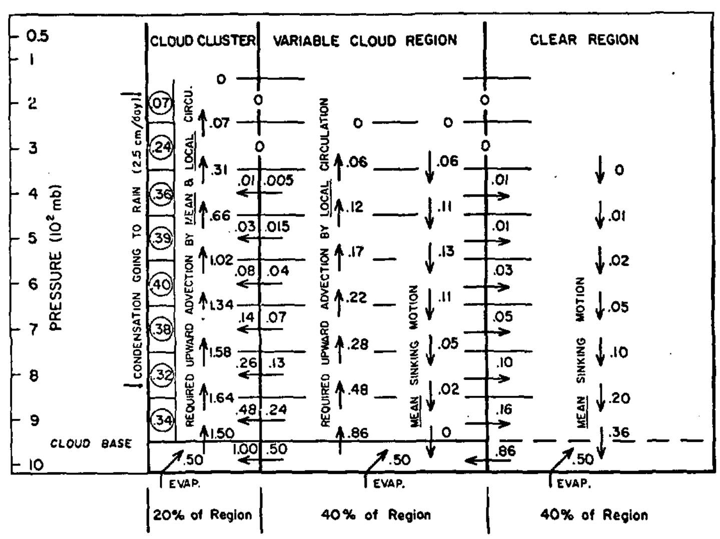

FIGURE 13.—Water-vapor budget ($g \cdot cm^{-2} \cdot day^{-1}$) of the tropical belt in summer. (The vapor advection per unit area into cluster is doubled because the variable cloud region is twice the area of the cluster region.)

need for upward vapor transport that cannot be accomplished by the synoptic scale or average upward circulation. One has to consider only the water vapor budget at each level from a physical or empirical point of view to learn what is required of the local vertical circulation. The required upward water-vapor transport ($g \cdot cm^{-2} day^{-1}$) by the subsynoptic circulation at any level $\hat{q}_{pi}$ must be given (for steady-state conditions) by adding the surface evaporation E_{sfc} to the integrated net inward horizontal water vapor advection q_H below level p_i and subtracting from these two quantities the vapor

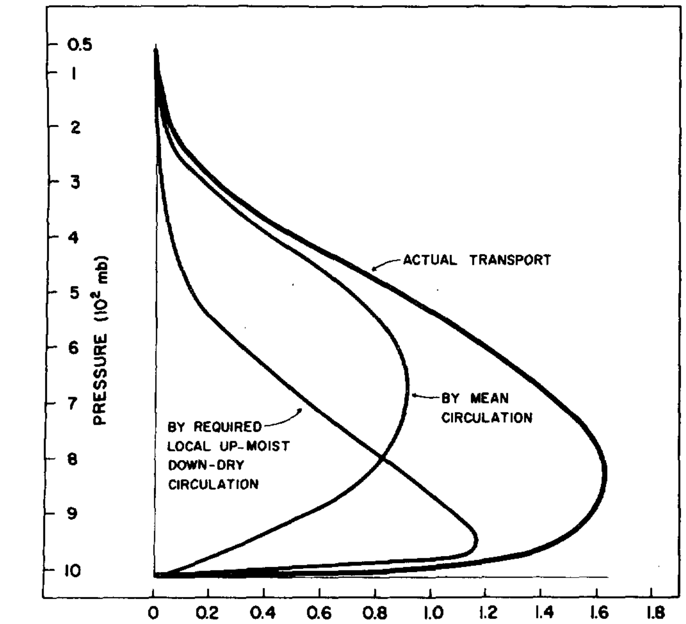

FIGURE 14.—Mean versus local circulation of upward transport of water vapor ($g \cdot cm^{-2} \cdot day^{-1}$) within the cloud cluster.

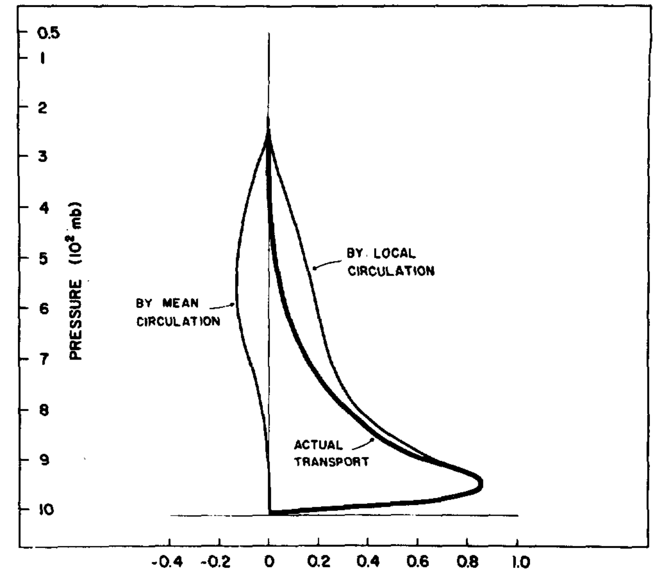

FIGURE 15.—Same as figure 14 except this is for the variable cloud region.

carried upward by the synoptic circulation $\bar{q}_p$ and the vapor that has been condensed to rain $\dot{R}$ below level p_i. The required upward vapor transport by the subsynoptic circulation at any level $\hat{q}_{pi}$ is thus given by

$$\left[\begin{array}{l}\text{upward}\\ \text{transport}\\ \text{of vapor}\\ \text{by local}\\ \text{circulation}\\ \text{at level } i\end{array}\right] = \left(\begin{array}{l}\text{evapora-}\\ \text{tion from}\\ \text{sfc}\end{array}\right) + \left[\begin{array}{l}\text{horizontal}\\ \text{vapor ad-}\\ \text{vection}\\ \text{from sfc}\\ \text{to level } i\end{array}\right] - \left[\begin{array}{l}\text{vapor}\\ \text{carried}\\ \text{by mean}\\ \text{circula-}\\ \text{tion at}\\ \text{level } i\end{array}\right] - \left[\begin{array}{l}\text{vapor}\\ \text{condensation}\\ \text{to rain below}\\ \text{level } i\end{array}\right],$$

$$\hat{q}_{pi} = E_{sfc} + q_H - \bar{q}_{pi} - \dot{R},$$

or

$$\hat{q}_{pi} = \text{evap.}_{sfc} + \int_{sfc}^{p_i} q_{Hi} \frac{\delta p}{g} - \bar{q}_{pi} - \int_{sfc}^{p_i} \left[\omega_m \frac{\partial q_s}{\partial p} + \omega_l \left(\frac{\partial q_s}{\partial p} - \frac{\partial q_e}{\partial p} \right) \right] \frac{\delta p}{g} \tag{2}$$

where g is gravity; i, the level of consideration; q_s, the saturated specific humidity; q_e, the specific humidity of the environment; ω_m, the mean or synoptic scale upward moist circulation; ω_l, the subsynoptic or local up-moist and compensating down-dry circulation; circumflex ($\hat{}$), the subsynoptic circulation; and overbar($\bar{}$), the synoptic circulation.

Figures 14–16 depict in more detail the water-vapor characteristics of the cloud cluster. Figure 14 shows that the subsynoptic circulation is most responsible for the upward transport of water vapor in the lower levels of the cloud cluster. The circulations have equal magnitude at 800 mb. The mean circulation dominates at upper levels. In the lower levels, the synoptic scale circulation in the cluster carries only a small fraction of the required upward transport of vapor necessary to produce the 2.5 cm/day mean rainfall. The rest of the upward vapor transport must be carried by the subsynoptic or local circulation.

At 950 mb, the synoptic scale vertical motion is only 20 mb/day. This results in an upward transport of only 0.34 $g \cdot cm^{-2} \cdot day^{-1}$ of water vapor. The water-vapor budget requires an upward transport of at least 1.5 $g \cdot cm^{-2} \cdot day^{-1}$. For the remaining upward vapor transport of 1.16 $g \cdot cm^{-2} \cdot day^{-1}$ to take place, there must be a compensating local or subgrid scale up-moist and down-dry vertical circulation of about 360 mb/day at 950 mb. At 850 and 700 mb, this local circulation must average about 280 and 185 mb/day, respectively. Although the net synoptic scale mass transport at each level by this local circulation is zero, the upward transport of vapor by the cumulus is greater than the downward transport of vapor between the cumulus. In this paper, we attempt to show the magnitude and fundamental importance of this local or subsynoptic scale vertical circulation to the tropical region vapor and energy budgets.

Figure 15 is a similar portrayal of the required upward water-vapor transport by the subsynotpic circulation for

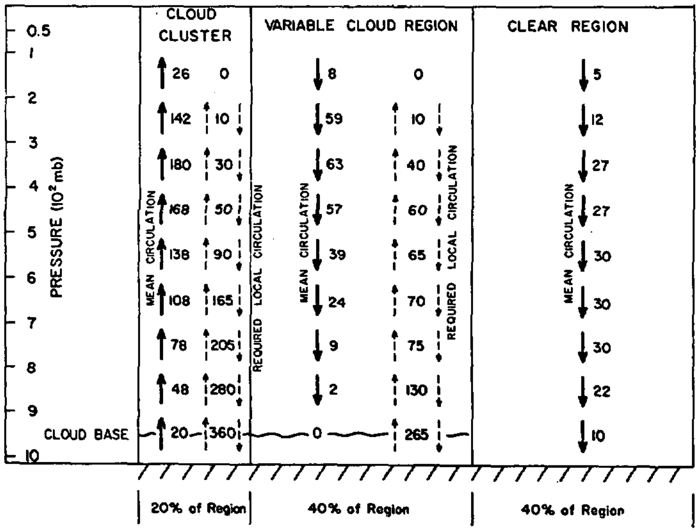

FIGURE 16.—Comparison of the mean and local vertical circulations (mb/day) in the various tropical regions in summer. The clear region has no local circulation.

the variable cloud region. In this region, the mean or synoptic scale circulation is downward. All the upward vapor transport must be accomplished by the local circulation.

Once the required upward vapor transport by the mass compensating or local circulation $\hat{q}_{v_l}$ has been specified at every level, the equal magnitude up-moist and down-dry circulation ω_{v_l} necessary to accomplish this upward vapor transport is determined at each level going upward from the cloud base using the equation

$$\omega_{v_l}=\frac{\hat{q}_{v_l}g}{q_s-q_e} \tag{3}$$

where q_{v_l} is in units of grams per square centimeters per day; the other symbols have been defined in eq (2).

In figure 16, we show the magnitude of this required local mass compensating up-moist and down-dry vertical circulation in millibars per day (dashed lines) for both the cluster and variable cloud regions and compare it with the mean or synoptic scale vertical motion in these regions. At cloud base (approx. 950 mb) these required local cirlations are no less than 360 and 265 mb/day, respectively. By contrast, the mean circulations are only 20 and 0 mb/day, respectively. Low-level convergence gives little indication of the actual up-and-down vertical circulation in operation at low levels.

Figure 17 portrays the contributions of the mean and of the local circulations to the production of rainfall. This figure contrasts the level-by-level upward vapor transports and the vapor released to rain by the mean and by the local circulations. Note that about three-fourths of the cluster condensation going to rain comes from the mean circulation. Since most of the condensation from the upward local circulation is re-evaporated into the compensating downward flow, the local circulation should be considered responsible for the vertical redistribution of vapor.

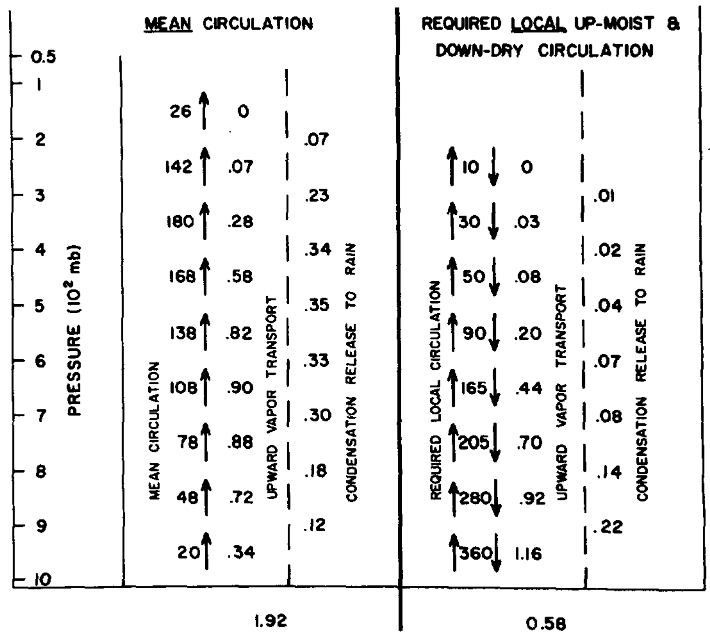

FIGURE 17.—Cloud cluster mean versus local circulation values (mb/day) of upward vapor transport (g·cm⁻²) and of condensation going to rain [g·cm⁻²·(100 mb)⁻¹]. At the bottom, the 1.92 and 0.58 (in g·cm⁻²·day⁻¹) are the resulting rainfall values.

Table 2 explicitly elists the vertical distribution of all parameters going into the typical cluster condensation, rainfall, and local vertical circulation determination as portrayed in the four preceding figures.

This required local or subsynoptic up-moist [4] and down-dry circulation thus is seen to be a fundamental component to the upward vertical transport of water vapor in both the cluster and in the variable cloud regions. At lower levels, it is responsible for most of the upward water-vapor transfer. Despite its importance, this mass balancing up-moist and down-dry local circulation cannot be resolved by synoptic analysis. It is true we reach a general understanding about its very large magnitude and probable important role in vertical transfer processes.

5. DETERMINATION OF INDIVIDUAL REGION ENERGY BUDGETS

Once the required regional net downward motion ω_d from both the mean and local circulations has been determined, the compressional warming, the water-vapor recycling, and the net cloud cooling can be obtained for each region from the need to maintain steady-state conditions. As previously discussed, interregional temperature advection influences are negligible. Thus at each level, the sensible temperature budget must be given as a balance between

$$\left(\begin{matrix}\text{sinking}\\ \text{warming}\end{matrix}\right)+\left(\begin{matrix}\text{cloud-induced}\\ \text{temperature changes}\end{matrix}\right)+\left(\begin{matrix}\text{net}\\ \text{radiation}\end{matrix}\right)=0$$

or (4)

$$\omega_d(\Gamma_d-\Gamma_e)+\left(\begin{matrix}\text{cloud sensible}\\ \text{temperature warming}\\ +\text{evaporation cooling}\end{matrix}\right)+R=0$$

[4] Up-moist or down-moist circulation means that the circulation occurs under water-vapor saturation.

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Table 2.—*Determination of cluster required local circulation*

Level	Mean circulation	Condensation to rain (mean and local circulation)			Moisture source (evaporation and horizontal vapor advection)	Moisture loss through condensation to rain	Total required upward vapor advection after subtraction of condensation to rain	Moisture carried by mean circulation	Required upward vapor advection at layer base by local circulation	Cluster saturated minus observed specific humidity	Required local up-moist and down-dry circulation	Level
(mb)	(mb/day)	[$g \cdot cm^{-2} \cdot (100\ mb)^{-1} \cdot day^{-1}$]			($g \cdot cm^{-2} \cdot day^{-1}$)	[$g\ cm^{-2} \cdot (100\ mb)^{-1} \cdot day^{-1}$]	($g \cdot cm^{-2} \cdot day^{-1}$)	($g \cdot cm^{-2} \cdot day^{-1}$)	($g \cdot cm^{-2} \cdot day^{-1}$)	(g/kg)	(mb/day)	(mb)
		•	†	‡	0.50§							
sfc	0				—⊙							
1000					1.00							
950	20	—	—	—	—⊕	—	1.50	0.34	1.16	3.2	360	950
900		0.12	0.22	0.34	.48	0.34						
850	48	—	—	—	—	—	1.64	.72	.92	3.3	280	850
800		.18	.14	.32	.26	.32						
750	78	—	—	—	—	—	1.58	.88	.70	3.4	205	750
700		.30	.08	.38	.14	.38						
650	108	—	—	—	—	—	1.34	.90	.44	2.7	165	650
600		.33	.07	.40	.08	.40						
550	138	—	—	—	—	—	1.02	.82	.20	2.0	90	550
500		.35	.04	.39	.04	.39						
450	168	—	—	—	—	—	0.66	.58	.08	1.7	50	450
400		.34	.02	.36	.00	.36						
350	180	—	—	—	—	—	.31	.28	.03	1.0	30	350
300		.23	.01	.24	.00	.24						
250	142	—	—	—	—	—	.07	.07	.002	0.2	10	250
200		.07	.00	.07	.00	.07						
150	26	—	—	—	—	—	.00	.00	—	0.0	—	150
100		.00	.00	.00		.00						
50	0	—	—	—		—						
Total		1.92	.58	2.50	2.50	2.50						

• Column for mean
† Column for local
‡ Column for total
§ Evaporation
⊙ Surface
⊕ Cumulus (CU) base

where Γ_d and Γ_e are the dry adiabatic and environmental lapse rates. López (1973*a*) has shown in his entire-life cumulus model that direct sensible temperature diffusion from cumuli to their environment is very small and, in comparison with the evaporational cooling induced by the cumuli, can be neglected. Most of the cloud-induced sensible temperature changes result from liquid water evaporation on the sides of the cumuli during their growth and throughout the interior of the cumuli as they die. This is portrayed pictorially in figures 18 and 19.

The sensible temperature budget of the summertime tropical belt is determined largely by the balance between sinking warming and by evaporational and radiational cooling. Thus

$$\text{sinking warming} \approx \text{evaporation} + \text{radiation}$$

or

$$\omega_d(\Gamma_d - \Gamma_e) \qquad L(\Delta q) \quad + \quad R \tag{5}$$

where Δ_q represents the amount of cumulus liquid water reconverted to vapor per unit mass and time on the sides of the clouds or within the clouds as they die. The energy balance given by eq (5) is portrayed in figure 20.

Throughout most of the summertime tropical belt, the sinking warming and evaporation cooling are much larger than the radiation cooling. Figures 21–23 portray the vertical profile of the terms of eq (5) for each of the three tropical regions. In all but the clear regions, the primary balance is between the sinking warming and the evaporation cooling with the radiation cooling of much less local importance. These cloud-induced cooling rates are much larger in comparison with the radiational cooling than was previously expected.

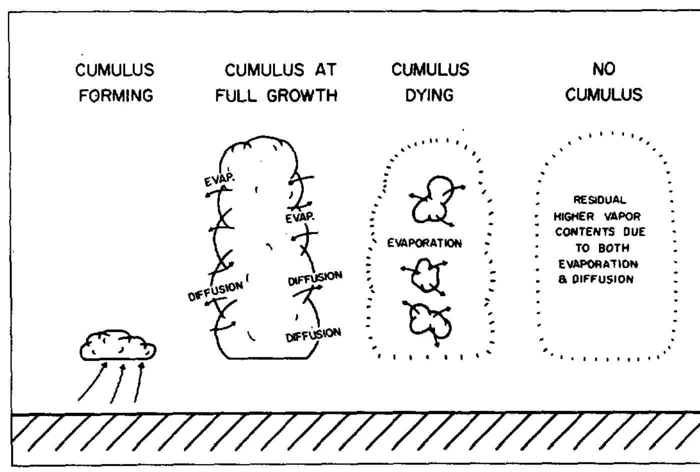

Figure 18.—Idealized picture of how the individual cumulus cools the atmosphere around it when it dies and how this cumulus raises the water-vapor level.

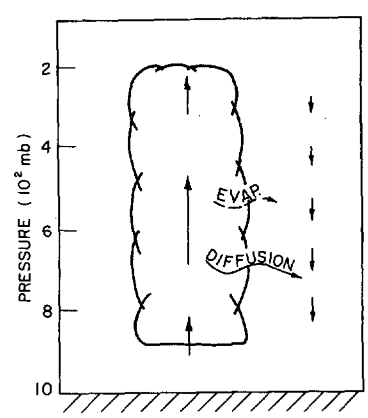

Figure 19.—Idealized picture of how the individual cumulus interacts with its environment.

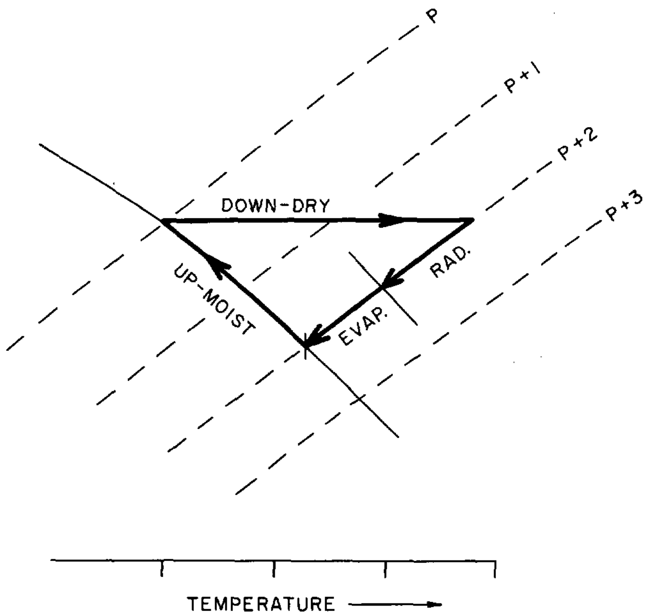

FIGURE 20.—Graphical portrayal of how an up-moist and down-dry vertical circulation can balance the radiation and evaporation energy losses.

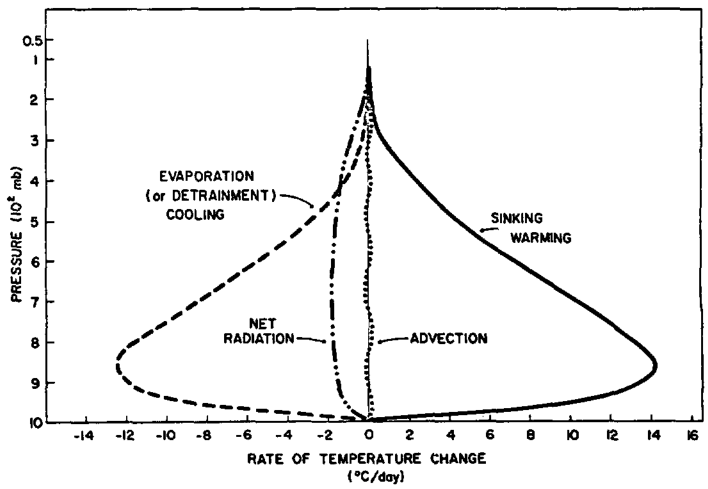

FIGURE 21.—Vertical distribution of required energy components in a steady-state cloud cluster.

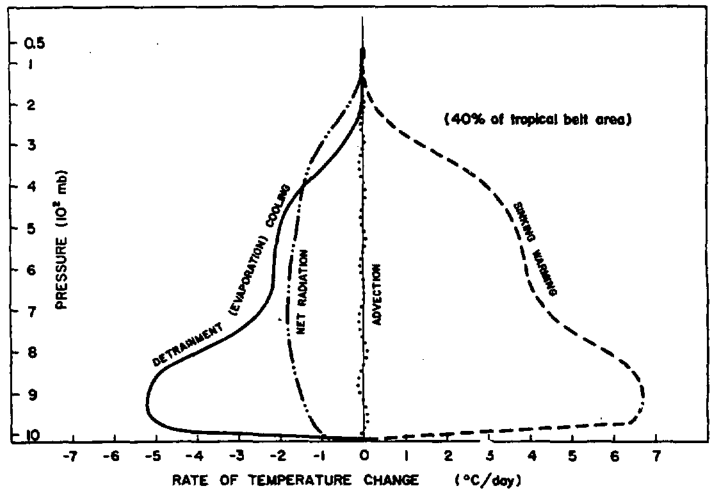

FIGURE 22.—Same as figure 21 except this is for the variable cloud region.

Vapor Replacement Rates.

Figure 24 shows the very large local recycling of water vapor for each of the tropical regions. Each region loses vapor through subsidence drying. For steady conditions to prevail, vapor must be advected into each region, or the individual clouds must furnish vapor either by (1) direct vapor diffusion out from the clouds or (2) advection or mixing of liquid water (to be evaporated) out from the clouds. Both of these processes are accomplished from the sides of the clouds or throughout the region occupied by the clouds as they die, as implied in figures 18 and 19.

The relative magnitude of these two processes of vapor replacement to the sinking motion is specified in figures 25 and 26 for the cluster and variable cloud regions. In the lower levels, most of the vapor replacement comes from direct diffusion or advection of vapor out from the cloud. At middle and upper levels, most of the vapor replacement comes from evaporation. This is a result of higher liquid water detrainment from the cumulus at upper levels and the lower upper level water-vapor contents. The percentage of water-vapor replacement by evaporation to total water-vapor replacement is shown on the right side of these figures. This ranges from 20 to 25 percent at 900 mb to 80 to 90 percent at 500 mb.

6. ACTUAL OR PROBABLE VERSUS REQUIRED VERTICAL CIRCULATIONS

In the preceding sections, we have discussed the required mean and local vertical circulation necessary to satisfy simultaneously mass, vapor, and energy needs. These mean and local required circulations must be continually in operation. This does not, however, specify all the subsynoptic vertical motion going on. Besides these required circulations, there is an additional subsynoptic circulation that does not influence the mass, vapor, or energy budgets and can only be roughly estimated. This extra mass-balancing local circulation is the additional up-dry and compensating down-dry circulation that accompanies forced upward motion (often associated with layered clouds) and the up-moist and compensating down-moist downdraft circulations associated with heavy rainfall. Any additional up-dry circulation not already specified must be mass balanced by a compensating down-dry motion. Similarly, any additional moist downdraft motion must be balanced by a mass compensating moist upward motion.

These additional mass-balancing circulations, for which magnitudes can only be roughly estimated, must be added to the already determined *required* up-moist and down-dry circulations to arrive at the actual or here-defined probable real vertical circulation. Figure 27 portrays the author's estimate of the actual vertical circulation in the cluster regions previously discussed. Seven classes of ver-

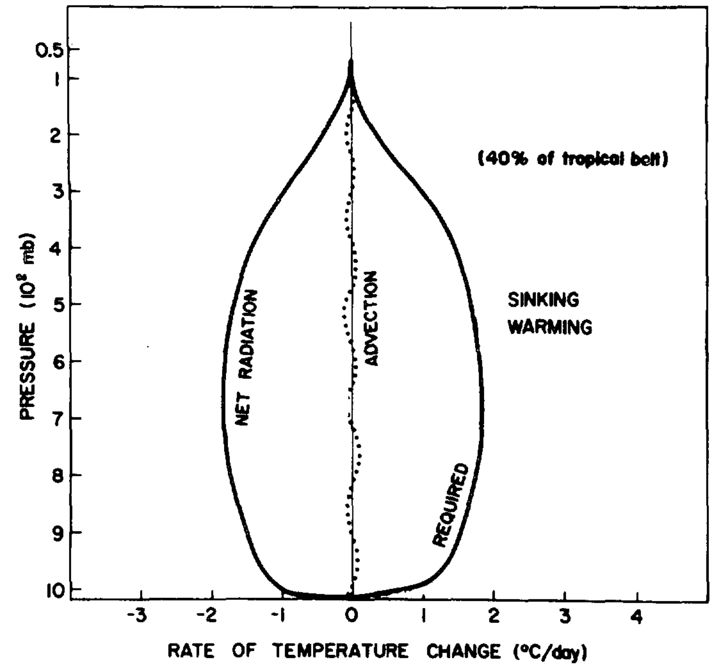

FIGURE 23.—Same as figure 21 except this is for the clear region.

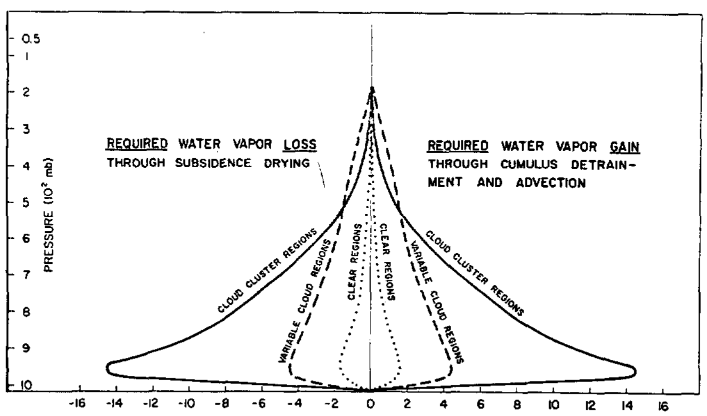

FIGURE 24.—Comparison of the required water-vapor gain and loss ($g \cdot kg^{-1} \cdot day^{-1}$) for each of the three tropical regions.

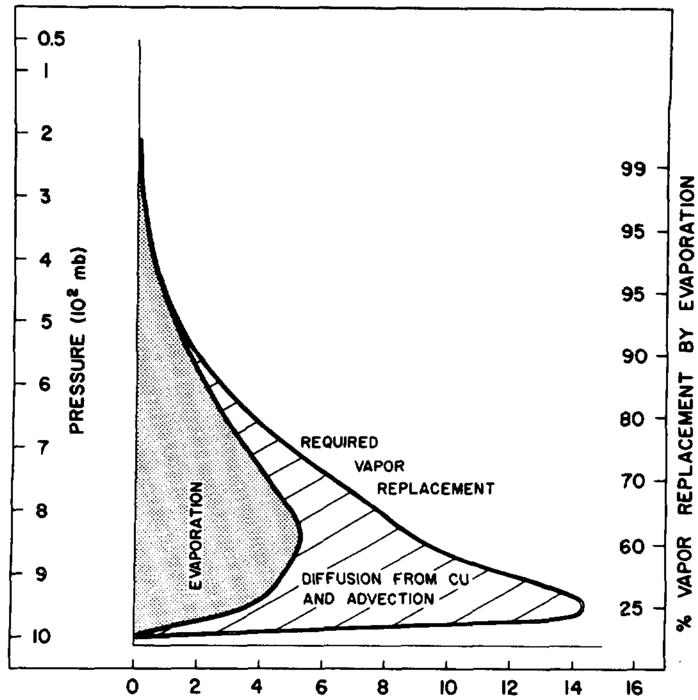

FIGURE 25.—Total required water-vapor replacement ($g \cdot kg^{-1}$ day^{-1}) to sinking motion to maintain cloud cluster conditions and that portion of the vapor replacement that comes from evaporation.

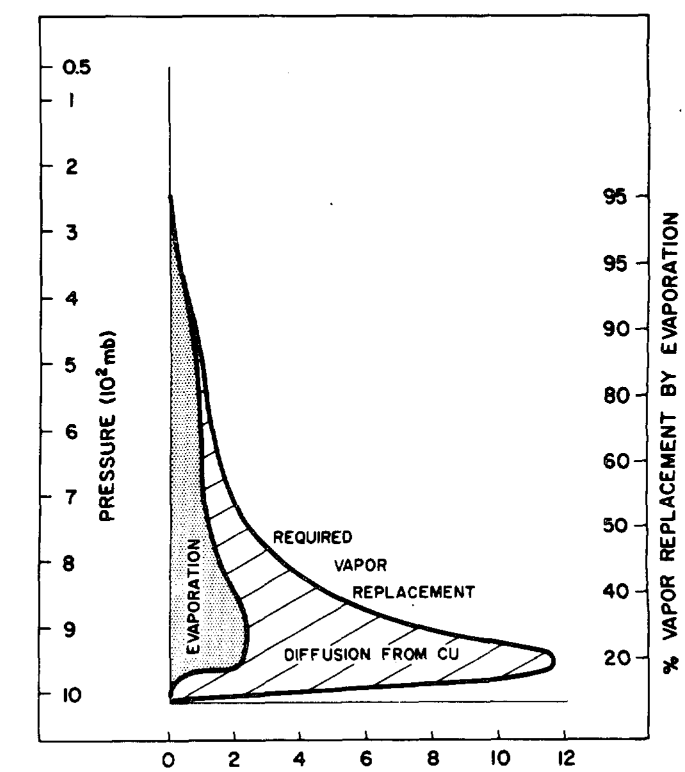

FIGURE 26.—Same as figure 25, except this is for the variable cloud region.

tical circulations are shown, four moist and three dry.[5] The moist or saturated motion has been shaded. These seven classes of vertical circulations are:

1. *Mean* upward motion as determined from the radiosonde cluster data (fig. 12). This motion has a maximum value of 180 mb/day at 350 mb. We assumed that the mean circulation goes up-moist.
2. *Required* local up-moist circulation with a maximum value of about 350 mb/day at 950 mb.
3. *Required* local down-dry circulation with maximum value of about 350 mb/day at 950 mb. Circulations 2 and 3 must balance each other.
4. *Probable* extra down-moist circulation as results with saturated downdrafts with rain. We assumed the maximum values in the lower troposphere to be about 200 mb/day. This estimate is quite subjective.
5. *Probable* extra moist updraft to balance the moist downdraft. Motions 4 and 5 are mutually balancing.
6. *Probable* extra up-dry motion from mesoscale or synoptic scale forced vertical motion as must be present to produce layered clouds.
7. *Probable* extra additional down-dry motion to balance the forced up-dry motion of motion 6. Motions 6 and 7 are mutually balancing.

[5] Although much of the real cluster vertical motion is not wholly moist or dry, it is more instructive to make a clear distinction between saturated and unsaturated vertical motion.

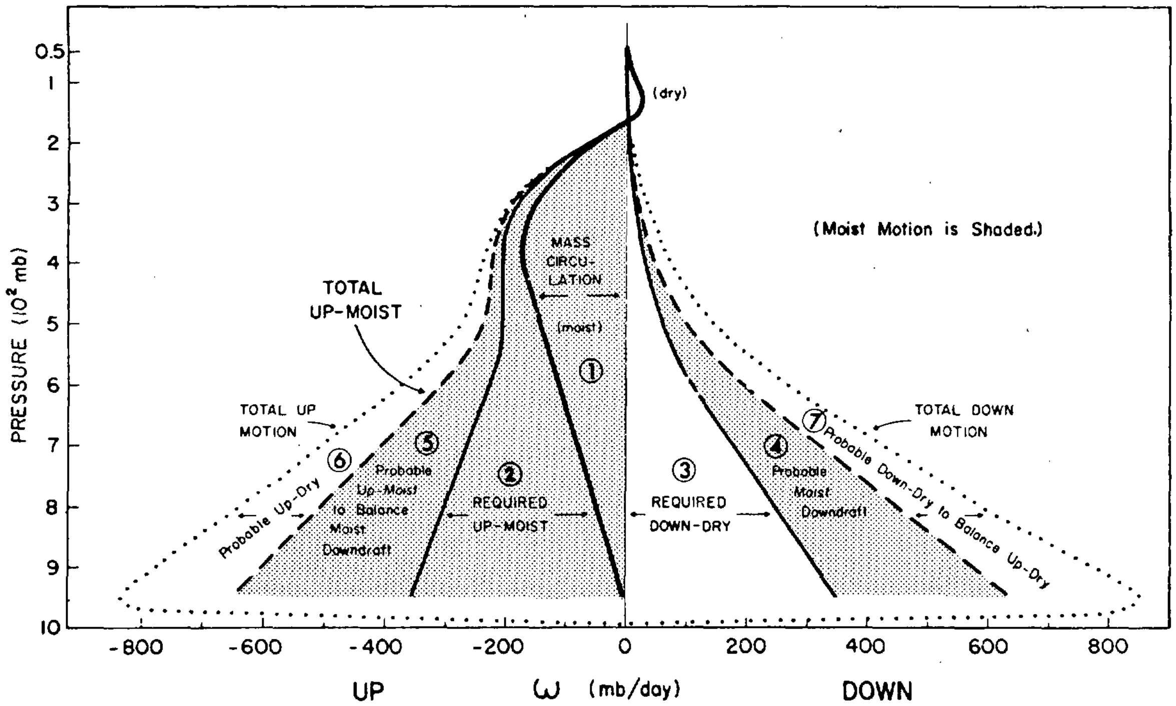

FIGURE 27.—Estimated complete vertical motion pattern of the cloud cluster region. The numbered individual vertical motion components are discussed in the text.

If this estimate of the typical cluster vertical circulation is not too inaccurate, then we are forced to accept the reality of a very large cluster recycling vertical circulation. In the lower half of the troposphere, this vertical circulation is from one to two orders of magnitude larger than the mean cluster upward motion. The implication of this very large local recycling circulation for vertical momentum and other dynamic influences may be substantial and requires careful consideration.

The divergence pattern implied by the vertical motion of figure 27 exactly fits the mean divergence pattern of figure 11. The cloud base to 400 mb cluster convergence occurs because the increase in net downward motion is larger than the increase of net upward motion.

Figure 28 shows the author's estimate of the actual or probable vertical circulations occurring in the variable cloud region. Again, one should note how large the actual up-and-down local vertical circulations are in comparison with the mean vertical motion.

In the clear regions, only a small extra or probable up-dry and down-dry circulation must be added to the mean circulation (fig. 29). This extra up-and-down dry circulation is thought to be due to boundary-layer turbulent mixing.

Such a large required up-and-down compensating vertical circulation was not expected at lower levels. Reed and Recker (1971) have also deduced from broad-scale considerations that a large lower tropospheric recycling circulation must be present, but their estimates are not as large. The implication of this large vertical circulation for the dynamics of the lower troposphere must be great. An explanation of the lack of vertical vorticity balance and the need to hypothesize substantial subsynoptic or cumulus vertical vorticity transfer, as has been discussed by Williams and Gray (1973), Reed and Recker (1971), Gray (1972b), and Holton and Colton (1972), are likely to be associated with this strong vertical circulation requirement.

This magnitude of cluster vertical circulation fits well the estimates in part II (López 1973c), which were derived independently from a cumulus modeling approach.

7. DISCUSSION

Large Vertical Recycling

The large subsynoptic vertical circulation or recycling of air that has been discussed is a function of the following two primary tropospheric processes:

1. The direct radiational cooling and precipitation warming of the troposphere.
2. The direct cooling nature of the cumulus clouds as they evaporate and die.

These two processes combine to produce substantial cooling of the atmosphere, especially in the lower troposphere. The net radiational cooling of the troposphere of about 1°C/day is well known and accepted. The idea of the cumulus cloud as "a direct cooling agent-an indirect warming agent" has yet to be fully appreciated or accepted. To understand how the troposphere balances its radiation loss, one must understand how condensation warms the atmosphere. Cumulus clouds (whether causing precipitation or not) do not directly act to warm the troposphere. In fact, we must conclude that cumulus clouds directly cool their environment. By what mechanism, then, do the cumulus clouds act to balance the tropospheric radiation loss and at the same time directly cool?

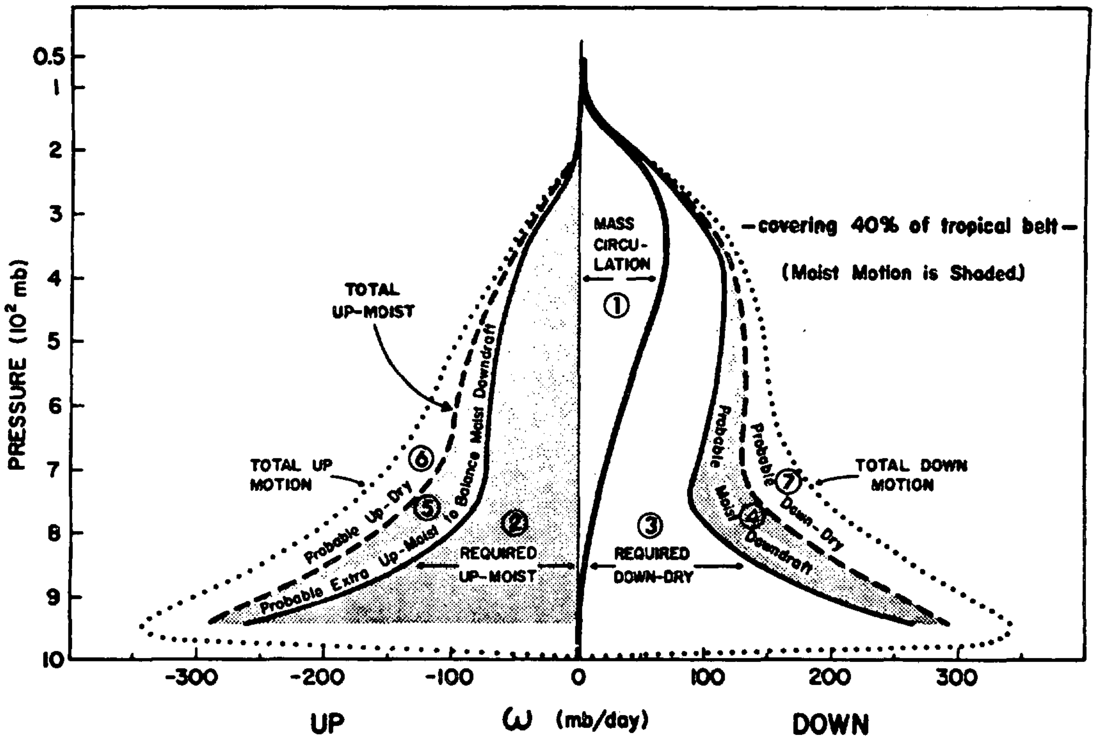

FIGURE 28.—Same as figure 27 except this is for the variable cloud region.

As paradoxical as this may appear, it is indeed the process that we believe is occurring. The latent heat released from cumulus clouds (both precipitating and nonprecipitating) goes primarily into potential energy gain and increasing the temperature of the rising parcel to that of the environmental temperature. The small extra (above environment) temperature increase of 1°–2°C of the rising parcel, which is required for buoyancy, does not warm the environment unless it directly mixes out from the cloud at a higher temperature. The rising parcel typically continues rising until it loses its buoyancy and temperature excess. It *then* mixes with the environment at a temperature little different (or even lower) than that of the environment. This does not warm the environment. Any diffusional or advective heat transports out from the rising (and warmer) cloud parcel are believed to be more than overcome by evaporation of the residual cloud liquid-water particles around the cumulus or evaporation of cloud particles as the cumulus dissipates. Individual small cumuli last only 5–10 min; cumulonimbi last only about 30–40 min. The residual liquid particles, which remain after the vertical motion in the cumulus has stopped, cool the environment in and around the dying cloud at a rate of 2.4°C for every gram per kilogram evaporated. Being an open system, the cumulus converts all its condensation heat to potential energy and exports this to the surrounding environment. Although rainfall may have occurred, there is typically no warming; instead, there is local cooling of the environment. This is not to say that the total effect of the condensation to the closed system of the globe or hemisphere has not been one of warming. It has. The dry adiabatic sinking at some other location (to satisfy mass balance) *has* more than compensated for the local cooling if there has

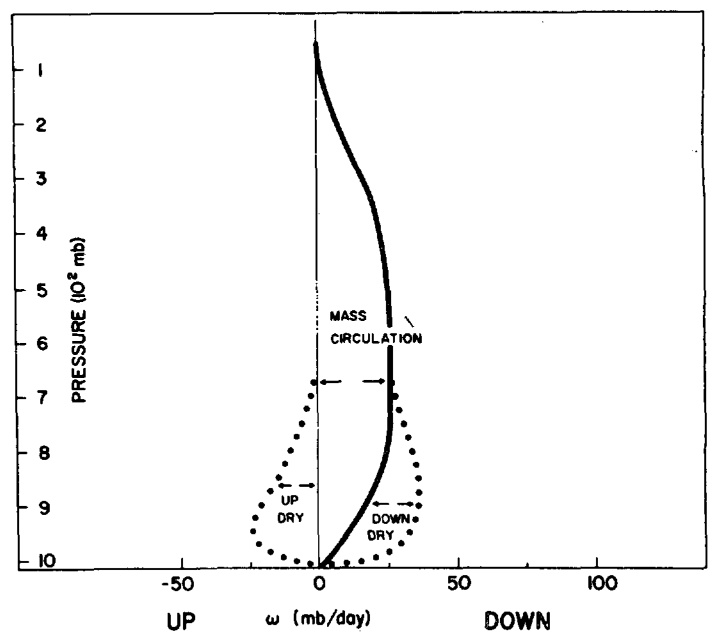

FIGURE 29.—Assumed total vertical motion in the clear regions.

been rainfall. In this way, the cumuli act in a delayed action sense. They produce a *local* sensible cooling, but a *global averaged* warming results.

The following evidence supports this physical explanation:

1. Direct sensible temperature decreases around and after cumulus convection has occurred, as reported by Kininmonth (1971) from data of the VEMHEX project of 1969.

2. The lower tropospheric cool nature of the raining tropical cluster and easterly wave, as substantiated by Riehl (1965) and many others since. Why does this condensation not warm the environment?

3. The lack of any tropospheric warming (or surface pressure falls) in an observational study of 1,000 Pacific trade-wind cloud clusters by Williams and Gray (1973). The average observed and calculated rainfall for these clusters was 2.5 cm/day for a 4° latitude square area. This 1500 $cal \cdot cm^{-2} \cdot day^{-1}$ condensation energy release led to no tropospheric warming, even though vertical wind shears and horizontal sensible temperature advective influences were small. Where did the heating go? Again, primarily to potential energy to be exported to the surrounding environment.

4. The direct cumulus cooling found by López (1973*a*) in his whole life cycle cloud model.

5. The maximum warming of the hurricane centers. Here, the highest temperatures are found in the center subsidence region. The region of maximum upward motion in the eye-wall convection area has lower temperatures. The eye temperature can only be explained by a compressional sinking motion in the eye, not by a diffusion of sensible temperature from the eye-wall cloud (Gray and Shea 1973).

This and other evidence indicates that we should view the cumulus precipitation warming process as one in which the cumulus initially cools the local environment but at the same time acts to warm the outer environment by compensating dry subsidence. The subsidence warming is larger than the local evaporation cooling and leads to a net large-scale tropospheric sensible temperature increase. Warming occurs at the place where the sinking motion exists. Local warming can, at times, occur if the compensating sinking motion from the cumulus takes place near the rising motion, as is the case with tropical storm genesis in regions of small vertical wind shear (Gray 1968, López 1968).

The primary sensible temperature input to the troposphere can only come from a continuous up-moist and down-dry vertical circulation, which acts to balance both the radiation and evaporation cooling as previously shown in figure 20. This sensible temperature gain, however, is paid for by a continuous water-vapor loss. The clear area down-dry portion of this vertical circulation continuously is losing vapor, which (for steady state) must be replaced by the cumulus. This is accomplished by vapor diffusion and advection from the variable cumulus regions.

A number of meteorologists now are thinking of the warming mechanism of the cumulus as resulting from compensating subsidence around the cumulus and in the more distant environment. The author recently has found general physical agreement on this idea in personal discussion and from the papers of Ludlam and Scorer (1953), López (1973*a*), Pearce and Riehl (1969), Ooyama (1971, 1972), Yanai et al. (1973), and Charney (1969). In addition, some researchers e.g., (López 1973*a*, Kininmonth 1971, Yanai et al. 1973) now are also accepting the idea of the cumulus cloud acting as a direct cooling agent. The earlier findings of Riehl (1945, 1965) and Elsberry (1966) on the lower tropospheric cooling of easterly waves might also be interpreted as generally supporting this point of view. Caution in literal interpretation of the "hot tower" hypothesis of Riehl and Malkus (1958, 1961) must be exercised. Their demonstration of the dominant role of vertical transports of mass occurring in selective cumulus or cumulonimbus towers is indeed correct, but these towers are (from the immediate environmental warming point of view) cool and not hot. They act to directly cool and not warm the environment. A cooling tower hypothesis does not, however, imply that the individual cumulus towers play any less fundamental role than that envisaged by Riehl and Malkus. The main problem of the cumulus acting as a direct heating rather than as a cooling source is that the required vertical circulation is reduced to unrealistic low values.

At a recent National Center for Atomospheric Research workshop on cumulus parameterization at Boulder, Colo.,[6] the general concensus was that a large recycling of the cloud-cluster atmosphere is required. This opinion is reflected in papers by López (1973*a*, 1973*c*), Yanai et al. (1973), Ooyama (1971, 1972), Rodenhaus (1972), and Zipser (1972). The author (Gray 1972*a*) has previously discussed this vertical recycling from a global point-of-view.

Role of Recycling in Generation of Individual Cumulus Elements

The general stability of the cloud-cluster lower layers and the need for a lifting mechanism to initiate parcel condensation until free convection is obtained is generally accepted. Over the oceans, this requires that substantial mechanical forcing be applied to initiate the buoyant elements. As discussed by López (1973*c*), it is impossible to generate cumulus over the oceans without cloud-base vertical velocities of 1–5 m/s. This demands convergence under the individual cumulus elements to be of the magnitude of approximately 5×10^{-3} s^{-1}, three orders of magnitude larger than the cluster synoptic convergence. Mean cluster boundary-layer convergences of approximately 5×10^{-6} s^{-1} do not permit typical cumulus area coverages of 1–5 percent to exist unless a substantial rain-induced downdraft mass flux into the boundary layer is occurring.

Relationship of Cluster Mass Recycling to CISK Mechanism

It is obvious that synoptic scale boundary layer induced frictional convergence from Ekman wind veering or the so-called "Conditional Instability of the Second Kind (CISK)" mechanism defined by Charney and Eliassen (1964) plays only a small *direct* role in the cluster upward mass transfers necessary to initiate the cumulus. This does not negate the fundamental need of the cluster for frictionally induced synoptic scale convergence, but it dictates that the primary direct function of the CISK process is probably to act as a mechanism for water vapor convergence. Unless the low levels of the cluster are continuously fed with water vapor, the rain processes would act to decrease the cluster vapor content. The number of rain-produced downdrafts and mass going into the boundary layer would be decreased, and cumulus convection could not be sustained.

The very high correlation of boundary-layer [7] relative

[6] July 10–14, 1972

[7] The terms boundary layer and subcloud layer are used synonymously. This refers to the layer from the surface to 950 mb.

vorticity with cumulus convection along frontal zones, along squall lines, on the cyclonic shearing side of the trade wind, and in tropical storms illustrates what a fundamental role the low-level frictional forcing plays in allowing the cumulus recycling mechanism to get going. The general lack of sustained cumulus convection in anticyclones and other regions of negative vorticity where weak lower level friction divergence is occurring is primarily due to the sinking induced drying of the environment. Vapor contents are not high enough to produce and sustain downdraft penetration into the boundary layer. It is then impossible to have sufficient mass convergence underneath new convective elements to sustain them over an area of 1–5 percent of the cluster. To understand the high correlation of cumulus convection with boundary-layer convergence and the CISK mechanism, one must fully appreciate the powerful influence of relative humidity on cumulus buoyancy and downdraft generation.

The cluster divergence profile of figure 11 is a result of the combined low-level frictional forcing mechanism and the cumulus feedback of the updrafts and downdrafts. Cluster convergence above the boundary layer is believed to be primarily a consequence of the cumulus. Middle-level convergence is necessary to feed the accelerating updrafts and downdrafts. Figure 27 shows that the 400–950-mb increase of the total downward motion is greater than the 950–400-mb decrease of the total upward motion. This middle-level convergence leads to additional water-vapor convergence. López (1973*b*) has recently indicated the powerful influence of high middle-level water-vapor content in initiating moist downdraft and specifying rainfall production.

Although middle-level and upper level broad-scale responses to the individual cumuli occur, this does not negate the fundamental role of the boundary layer in dictating these responses. The CISK mechanism should be viewed not as a direct cumulus-producing process, for in this it is much too weak, but instead as a required ingredient for recycling. As far as cumulus convection is concerned, it appears that the most important aspect of the CISK mechanism is not the magnitude of the direct boundary-layer mass convergence but rather the associated water-vapor convergence.

Possible Importance of Recycling for Vertical Momentum Transfer

Since the cluster sinking warming and the evaporation cooling largely balance each other and since the up-and-down recycling mass largely balance, one might feel that it is unnecessary to deal with the recycling mechanism directly but only with its net influence to the synoptic scale motion. From the cluster mass and energy budgets, this may be a reasonable hope; however, for the water-vapor and momentum budgets, it is probably not. It seems likely that there would be significant sub-synoptic water-vapor and vertical momentum rearrangements from this vertical circulation that cannot be directly specified in terms of the broad-scale flow. If this is the case, then the recycling process has important vapor and momentum feedback influences that have to be separately parameterized from, as yet, still-unknown physical processes.

This paper has shown how it is possible to arrive at an estimate of the vertical recycling in the tropical atmosphere from direct physical reasoning and without extensive mathematical analysis. It is hoped that this type of physical and graphical analysis can be used by others attempting to deal with atmospheric moist processes.

ACKNOWLEDGMENTS

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